Columbia University in the City of New York

LAMONT GEOLOGICAL OBSERVATORY

PALISADES, NEW YORK

OCEANOGRAPHIC WORKSHOP ' 1964

Edited by

Arnold L. Gordon

Technical Report No. CU-12-64 to the Atomic Energy Commission
Contract AT(30-1)2663

September, 1964

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(Columbia University)
Palisades, N. Y.

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In the period from June 8 thru June 12, 1964, the physical oceanography department at Lamont Geological Observatory held an Oceanographic Workshop. The object of this week of lectures and discussions was to acquaint people who had only an introductory knowledge of oceanography with the skills, methods and instruments, as well as the current problems of physical and chemical oceanography.

The term, "Workshop", was used rather than "Seminar" because the philosophy of these meetings was to give the participant a feeling for practical oceanography in the field and in the laboratory. He was able to inspect oceanographic instruments and discuss their operation with people who actually used them in the field. He was able to learn about procedures and methods of oceanography, their meaning and limitations, from people who have applied them to oceanographic data. In short, the Oceanographic Workshop is a supplement to an introductory oceanography course.

The original impetus for the Workshop came from Dr. T. Ichiye, and through his interest and that of Dr. G. Wüst and Dr. M. Ewing, the Workshop was organized and carried out successfully. The people who gave lectures at the Workshop are commended for their fine job of producing clear, informative talks, which were rewarded by the interest of the listeners and favorable results of the Workshop. I wish to thank all those who contributed to the sessions and made this report possible.

"Oceanographic Workshop, 1964" is a series of outlines which cover the lectures given on each of the five days. The lectures were from one and one-half to two hours long, generally followed by one half hour of discussion, though comments were accepted during the lecture period also. Slides, models and instruments were used during the lecture to clarify points or demonstrate procedure. Occasionally, the early afternoons were devoted to more detailed discussion between the speaker and several of the participants.

The scope of the Workshop was broad. As many topics were touched upon as possible in the alloted time with a few of the more important aspects treated in a more complete manner. The more detailed part of each of the lectures reflected the recent research work of the speakers.

The interest generated by this Workshop has indicated that future Workshops would be welcome. In these future sessions, the lectures would be more specialized; they would expand and add to the contents of "Oceanographic Workshop, 1964" rather than be a duplication. The outlines of this report will be open to discussion as well as the new material. The additional material covered at future

Workshops will be added as a supplement to the preceding one, and so the Workshop reports will grow and gradually become more complete. As an example of future specialization, there would be an instrument section which would devote the day to the new continuous in situ recorders. Additional sections will be added, such as the processing of oceanographic data by computer. In this way, the Oceanographic Workshop will expand in the same direction as oceanographic research. The time and place of future sessions will be announced.

The following is a list of institutes which were represented at the first session:

New York University (Dept. of Meteorology and Oceanography)

Rutgers University (Dept. of Geology)

U. S. Merchant Marine Academy

City College (Dept. of Geology and Dept. of Meteorology)

Hudson Laboratories

Sandy Hook Marine Laboratory

Columbia University (Zoology Dept.)

Lamont Geological Observatory

Arnold L. Gordon
September, 1964

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* Since the lecture Dr. Sackett left Lamont and did not prepare an outline for this report. Dr. Thurber was away at the time this report was compiled.

General Introduction to Instrumentation

Robert D. Gerard

Today the basic tools for the study of ocean waters beneath the mixed layer are not substantially different from those used fifty years ago. The implications of this statement are numerous. There is good testimony that the early methods and tools were soundly conceived. It bears out the urgent need for new methods, which in the past have developed faster than the accumulation of data from the oceans. Today our concept of the general circulation of the oceans is based mainly upon indirect observations of the density distribution and upon the distribution of the conservative properties of observed temperature and salinity. New electrical and electronic methods for continuous measurement of temperature, salinity, pressure and dissolved oxygen are now in the early stages of development and will certainly become the standard methods for future data collection. However, at the present time, the older methods of measurement at discrete depths of individual samples must accompany the continuous measurements in order to control the newer techniques.

A vertical cross-section of the typical ocean reveals a horizontal to vertical ratio, approximately 1000:1. This may be represented in proper scale by a chalk line across a blackboard ten feet long. It is remarkable, therefore, that in spite of this two-dimensional aspect, the oceans nevertheless exhibit a high degree of stratification. The physical oceanographer is concerned with the movement of water within this stratified system. The most satisfactory means to determine ocean currents will depend upon the physical scale and time scale which is involved. For the present, discussion of the methods of ocean current determination may be considered under the two major categories: 1) indirect measurements and 2) direct measurements. The details of these two approaches are covered in the Workshop outlines, and the reader is referred to these outlines. However, a few additional remarks can be made.

The indirect method can be no better than the density determinations which, in turn, are limited by the limits of error of the instruments employed. For the determination of density, precise determination of temperature, depth and salinity are required. For the past forty years, measurements of temperature and depth have been made at standard hydrographic stations using deep-sea reversing thermometers. In recent years almost everyone has agreed that the mercury thermometers to determine temperature and depth in the ocean will be replaced by electrical methods. Surprisingly, by far the greatest number of measurements being taken today are still taken with the mercury thermometers. It is surprising because the means have been available for a number of years to take precise electrical measurements of temperature in the ocean. The first instrument to regularly obtain precision temperature measurements throughout the water column in the deep ocean has been the Lamont thermograd.

Ironically, these temperature measurements in the water column are obtained as a bonus and are incidental to the major job for which the thermograd was designed, that of obtaining thermal gradient measurements in the sediments of the ocean bottom. Probably the reason that other instruments for electrical measurement of temperature are only now coming into use is because there has not been until very recently a satisfactory electronic method for in situ measurement of salinity, so that little effort was spent to develop devices for temperature, depth measurements alone. Today, however, two prototype instruments are available for purchase, which will provide a deck read-out of the temperature, depth and salinity from continuous in situ sensors. Both use an electrical resistance thermal element with a time constant of less than 1/2 second, and the temperature accuracies are listed variously between .05°C and .02°C.

From the thermograd records on hand at Lamont, it would appear that many parts of the ocean exhibit a vertical temperature profile that contains many heretofore unobserved reversals and changes of gradient, indicating thermal structure on a scale that is not observable with the standard oceanographic casts.

Surface ocean currents induced by wind stresses and horizontal pressure differences may reach more than 5 kts., while the average is less than 1 kt. Tidal currents are poorly known in the deep sea, but are believed to increase near land. Most surface current data has been gained from the summation of masses of ships' log information, where seasonal averages are prepared in a statistical manner. Climatological type current charts are available from various national hydrographic agencies. For more exact measurement of surface and sub-surface currents a great variety of devices have been used. The nature of measurements may be classified in two categories: Eulerian methods where velocity of flow past a fixed point is measured as a function of depth and time; and Lagrangian methods where the trajectories of tag particles or drifting objects are plotted with respect to time. In all of these methods, measurements of velocity, direction and depth and time are required.

The following section of the Workshop by Robert Sexton touches on the various problems of instrumentation. Some certain aspects of instrumentation are developing so rapidly, it would be impossible to have a thorough discussion in one day. The problems of salinometry are not dealt with in great detail but are reserved for discussion at future Workshops. The most rapid development is in the perfection of continuous in situ recording instruments. The data collected by these instruments will undoubtedly help the theoretical oceanographer solve many of his problems, but also create many new ones. (i.e. an explanation of micro-structure).

Oceanographic Instruments
Hydrographic work at sea

Robert K, Sexton
Monday, June 8, 1964

Introduction

This section of the oceanographic workshop is concerned with some selected instruments used in determining ocean currents, with considerations of the techniques involved in their use at sea, and with their accuracies. The subject is approached from the point of view of one who is interested in the instruments as tools for collecting oceanographic data and not from the mechanical or electrical point of view. The list of instruments to be discussed is far from complete. Rather, the emphasis is placed on the instruments which are most commonly used by oceanographers and which demonstrate the various approaches to the problem of measuring ocean currents. The literature which contains the information discussed here is extensive or widely scattered, and is many times written in general terms which makes the task of measuring water flow deceptively simple. The aim of this section is to bring together in one place an outline which points out difficulties in technique which are not at first obvious.

Almost ten years ago Böhnecke wrote: "the subject of current measurements has kept the oceanographer busy for more than a hundred years without having found...(this must be honestly admitted) ... an entirely satisfactory solution" (Böhnecke, 1955). Although the decade since this was written has seen a major increase in oceanographic effort, the statement still remains valid. The measurement of ocean currents is a complex problem, which is many times not clearly defined due to our ignorance of the relative magnitudes of the mean steady state component of flow to the time dependant component, and the time and space scale of the numerous turbulent eddies which are contained in the water flow. After we have more data, these problems will be clarified and the method for direct current measurements will be a more effective one.

There are two major approaches to measuring ocean currents, the indirect method and the direct method. These will be discussed separately.

The indirect method or dynamical method - This method is discussed by Arnold Gordon in the following section of the workshop. In general, it can be said that the dynamical method is the determination of water flow from the internal field of mass.

I. The Hydrographic station

The internal field of mass is calculated from the temperature and salinity data taken at various depths at the standard hydrographic station. The hydro-station is the classical approach to physical oceanography and is, to date, chiefly responsible for our knowledge of the distribution of the properties in the sea. The hydro-station is essentially the measurement

of temperature and the sampling of the sea water at various depths. Although this is one of the earliest techniques in oceanography, one which has been well standardized, the variation in the quality of data taken by different observers is striking.

II. Parameters measured for dynamic calculations. Density is a function of temperature, salinity and depth. The anomaly of specific volume is broken into three major components

$$\delta = \Delta_{S,t} + \delta_{S,P} + \delta_{t,P}$$

Thus, these three parameters must be known: Temperature (t), Salinity (S), Depth (p)

A) Temperature in degrees centigrade is measured with the protected reversing thermometer. Detailed descriptions of this instrument may be found in any standard textbook. It is essentially a double-ended mercury thermometer, which, when turned through 180 degrees, records the temperature at reversal by the separation of the mercury column. The mercury "locked" in the calibrated capillary tube expands when brought aboard the ship, since the air temperature is usually warmer than the sea water temperature at thermometer reversal. Contraction would occur if air temperature were colder. Because of this change in volume; of the mercury and of scale imperfections, a series of corrections must be made in the temperature which is ready aboard the ship to arrive at the actual temperature of the water at reversal.

To facilitate these corrections, a small auxiliary thermometer is paired with the main thermometer. The auxiliary thermometer gives the temperature at the time of reading. The paired thermometers are enclosed in a heavy-walled glass case as a protection against the pressure of great depths. Thermal conductivity is achieved through a mercury bath inside the pressure case. The most precise correction is given by:

$$\Delta T = \frac{(T' - t)\ (T + V_o)}{K - 1/2\ (T' - t) - (T' + V_o)}$$

when ΔT is the correction to be added algebraically to the uncorrected reading T'; t is the temperature at which the instrument is read (corrected auxiliary thermometer reading)*; V_o is the volume of the small bulb and the capillary up to the 0° mark and K is a constant depending on the type of glass used (K = 6100 or 6300). The calibration correction of the main thermometer, I, which depends on the value of T, must be added to ΔT to obtain the final corrected temperature.

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* The auxiliary thermometer reading is also corrected for scale imperfections.

Keyte (1964) has demonstrated that the formulae given in The Oceans by Sverdrup, Johnson and Fleming and in Physical Oceanography by Defant are incorrect. They should read

$$\Delta T = \frac{(T' - I)\ (T' + V_o)}{K} \left[I + \frac{T' - t) + (T' + V_o)}{K} \right]$$

The texts' omission is the second plus in the braces and can lead to an error of as much as 24%.

B) Depth can be found by the unprotected-protected reversing thermometer pair. The unprotected reversing thermometer is essentially the same as the protected thermometer, except that the glass case is open at the bottom to the effects of pressure. The difference in readings between the corrected protected thermometer and unprotected thermometer gives a measure of pressure and, therefore, of depth. Corrections must also be applied to the unprotected thermometer. In its most precise form, this is given by:

$$\Delta T_u = \frac{(Tw - t)\ (T' + V_o)}{K - 1/2\ (Tw - t)}$$

where Δ Tu is the correction to be added algebraically to the main unprotected thermometer reading; Tw is the corrected reading of the protected thermometer. A correction, i, must also be added to the unprotected auxilliary thermometer. Depth is given by the formula:

$$D = \frac{Tu - Tw}{Q \cdot Pm} \times 10$$

where D equals depth of reversal in meters, Q is the pressure factor in kilograms per square centimeters and Pm is the average density in situ.

C) Salinity has been found in the past by a chemical titration; this chemical method is occasionally used today. The ions belonging to the Halogen group is precipitated with a silver nitrate solution. The relation between the Halogen ions (chlorinity) and other ions in the sea water solution was found to be a very constant ratio in the 77 CHALLENGER samples which Dittman studied in 1884. From Dittman's study, Knudsen developed the following imperical relation which can be used to convert chlorinity to salinity.

$$S = 0.030 + 1.8050 (Cl\ ‰)$$

Since this relation is based on only a few quantitative analyses

of sea water and indications are that for various types of sea water (as shelf water) the formulae may be incorrect.

D) Salinity is now found by electrical methods. Conductivity is measured and a careful temperature compensation correction is made. The relationship between electrical conductivity to salinity is questionable and the problem is at present under consideration by an international committee.

III. Water sampling devices used at a hydrographic station

A) Nansen type bottle. Developed in the early part of the century, it remains essentially unchanged. The metal bottle is fastened to the hydrographic wire and lowered to a predetermined depth, where it is reversed with a messenger. In the process of reversing, it captures a water sample which is brought back to the surface for analysis, and reverses the thermometers attached to the Nansen bottle. These bottles require periodic maintenance to assure that their valves remain water-tight. At the start of an expedition, the bottles should be disassembled and cleaned. Occasionally, the valves require reseating. A check should also be made on the condition of the teflon lining of the bottle, the function of which is to inhibit any reaction of the water sample with the metal parts of the bottle. This is particularly important in taking samples for dissolved oxygen content. The water collected by the Nansen bottle must supply the necessary amount of sample for the routine chemical analysis aboard the ship.

B) There are other samplers which trap the sea water when hit by a messenger, not all of which reverse. The Van Doren bottle is used occasionally when only a water sample and not temperature is needed; this bottle does not reverse when hit by the messenger, but clamp two rubber plungers on both ends of the bottle. They are light and inexpensive, though not so durable as the Nansen bottle or as easy to use.

C) Other types of samplers use propellers to lock the bottle (Sigsbee and Buchanan) or the hydrostatic pressure (Spilhaus).

IV. Errors involved in measuring the temperature and salinity

A) In the reversing thermometer

1) A careful thermometer calibration is needed to correct for errors in the scale markings. As a general procedure, each new thermometer is tested first by the manufacturer and then by the

U. S. Navy Oceanographic Office Calibration Facility. The Navy Oceanographic Office tests 5 points over the scale of each of the main and auxiliary thermometers. The V_o or the volume of the small bulb and capillary up to the 0° mark, is also checked. In the case of the unprotected thermometer, about 5 Q values for different depths are established. These are assumed to be linear -- a somewhat dangerous assumption. The thermometers are then tested by periodic ice-point determinations. These last tests are necessary because the aging of the glass, particularly over the first year of the thermometer's life, may cause an alteration in the correction factor, I. Generally, a distilled-water ice bath in an insulated vessel is used for the ice-point determinations.

2) Malfunctioning of the reversing thermometers

This may be considered to occur when the mercury breaks either at the wrong place, that is, some other place than precisely at the appendix, or when the mercury does not break at all.

3) Incorrect readings or correction of readings

The procedure for reading the thermometer should be carefully considered. Here at Lamont we remove the thermometer frames from the Nansen bottles after they have been brought back aboard the ship and put them into a fish tank filled with water. The purpose of this is to bring all the auxiliary thermometers to the same temperature and to hold them there for a period of time sufficient for two people to read them and to compare their readings. A magnifying lens with a hair line is used to read the thermometer through the glass plate of the fish tank. Every attempt is made to avoid paralax. In the case where two people read different temperatures on a particular thermometer, the thermometer is reread until full agreement is reached.

4)Approximate values for the various kinds of errors

a: Errors of reading - the accuracy of the reading can safely be assumed to lie within the limits ± .01°

b: Corrections arising from reduction errors (mathematical) no value can be assigned. If the errors are large, they are easily spotted.

If small, they may slip by unnoticed.

c: Correction errors arising from change of zero point - This usually ia a minus correction, because the aging of the glass tends to cause the contraction of the bulb, which causes a rise of the zero point. In the case of the 1/20 (.05) thermometers, this may amount to .015°C; and in the case of the 1/10 (.1) thermometers, it may amount to .02°C, the corresponding maximum values being about .035°C and .04°C. The errors of the breaking-off device are seldom more than ±.02 or ±.03°C.

d: The overall error of reversing thermometers is generally given as ±.01°C. However, it should be remembered that this is an idealized error.

B) Types of errors in measuring depth with reversing thermometers (Thermometric Depth) Since the difference in the readings of paired protected and unprotected thermometers is used to determine thermometric depth, all the errors referred to in the previous section also apply here. In addition, two more sources of error must be considered:

1) The Q factor must be known accurately and at many pressures so interpolation errors are reduced.

2) The value of ρ_m. Wüst has worked out mean values of ρ_m for the Atlantic Ocean. These values can, apparently, also be used in the Pacific and Indian Oceans. But, he cautions: "for secondary seas with large salinity anomalies, it is preferable to work out special tables of mean density. (Wüst, 1932)).

3) Using carefully calibrated thermometers graduated in .1°C, thermometric depth determinations are accurate to $\sim$0.5%. If the graduations are every .2°C, the error could be larger.

C) The errors in measuring salinity are not well known. Besides the usual machine (salinometer) errors, there are uncertainties in one's knowledge of the relationship of chlorinity to salinity and conductivity to chlorinity (salinity).

1) Cox (1963) has computed the standard deviation for various salinometers and finds that with a good operator, ± .01% standard deviation is attainable. He reports: "It is not difficult to find titrated results (Knudsen titration) with a standard deviation of .04% or .06% ."

2) In general, oceanographers say that the chemical titration method is "accurate" to ± 0.01 o/oo and the electrical methods to ± 0.003 - 0.005 o/oo. The salinometer we use here at Lamont is the Hytech Inductive Salinometer and Conductivity Meter. This ideally permits determinations of salinity to ± .003 o/oo.

3) A few operational techniques are worthy of mention. Since temperature differences are compensated for within the salinometer, it is required that the samples to be measured and the standard sea water be within 3°C of the room temperature before starting the measurements, and also, that the standard sea water must be within 2°C of the sample. This has the practical consequence that samples must be stored in the same room with the salinometer for about twelve hours before they may be run and that the room be one in which there is relatively little temperature fluctuation. Ideally, it should be an air-conditioned room, particularly for work in the tropics. Measurements on a batch of samples should always be preceded and followed by measurements of the standard water. It is considered good practice to check the machine by running the standard water about once every hour. Other operational techniques are:

a: the rinsing of the cell. The cell should be rinsed before each new sample is run.

b: the speed with which the measurement is made. If the sample is left in the salinometer for more than a few seconds, heating beyond the compensation ability of the machine may result in small errors.

c: periodic maintenance and cleaning of this cell is necessary. It is only with meticulous care that the accuracies that have been given will, in fact, be met.

D) Errors may occur due to poor sampling techniques. In order to insure that the samples which are to be measured are not contaminated, the following precautions should be observed:

1) The Nansen bottles should be water tight.

2) When the water is drawn from the Nansen bottle, at least two rinsings of the bottle should be made before

the final sample is drawn. Fill the bottle about a quarter full, put the cap on, shake it, take the cap off, pour the water out over the stopper. Repeat this. Finally, to draw the sample, hold the bottle carefully so that the stream of water coming from the Nansen bottle spigot does not touch the outside of the sample bottle, but flows directly in. Since many samples will be well below laboratory temperature when they are drawn, the sample bottle should only be filled up to the shoulder, leaving a small air space for the thermal expansion as the sample warms to room temperature. One precaution to be observed in drawing salinity samples is to keep the drops of water from the exterior of the Nansen bottle from contaminating the sample. Large quantities of water accumulate on the base of the thermometer frame above the stopcock. The best way to remove this is to blow it away. One must also be sure there are no drops of water on the stopcock itself.

V. In situ measurements of salinity, temperature and depth

A) In the last few years considerable progress has been made in measuring electrically the parameters of temperature, salinity and depth. Although the new instruments have not achieved the high accuracy of the reversing thermometers and the laboratory salinometer, they have a number of distinct advantages. Instead of about a dozen discrete measurements of temperature and salinity, as with the classical hydrographic station, one may obtain hundreds or even thousands of measurements in the same vertical column.

B) A couple of these instruments are:

1. The Industricl Instruments salinity, temperature measuring device (no depth transducer)

Accuracy:	Temp.	± .5°C
	Sal.	± .3 o/oo
	Depth	400 ft.

This instrument is good for crude survey work when a large change in temperature and salinity can be expected. It provides no permanent record.

2. The Hytech in situ temperature/salinity/depth recorder

Accuracy:	Temp.	±.05°C
	Sal.	±.03 o/oo
	Depth limit at present is 3000 meters within ± .25%.	

This instrument has enormous potential. The data is presented as a plot of temperature versus depth and salinity versus depth. A small adjustment allows the direct plotting of the T/S diagram. This instrument also lends itself to recording on magnetic tape and to computer processing.

Direct measurement of ocean currents - Direct measurement of water flow is not new in oceanography. The methods have developed slowly, but in recent years more attention has been given to the problems; direct methods are still in an experimental stage. The earliest useful measurement of currents were performed by Pillsbury in 1886 aboard the BLAKE in the Florida Straits. His results and results of geostrophic calculations (Wüst, 1924) compare very well. With more and more attention given to smallscale and transient features in the oceans, technique of the direct methods is an important problem to be solved, and is one of the most difficult tasks of physical oceanography.

I. Lagrangian Methods (the method of following a tagged particle)

A) Ship's drift is the basis for most existing surface current charts of the ocean. These are inadequate because of

1) the inaccuracies of navigation (± 1 sea mile)

2) the unknown effect of the wind acting on the ship (which is different for each ship structure) and the water flow on ship's kiel.

B) The Swallow float or the neutrally-buoyant float is ballasted so that it will sink to a particular density level in the sea and remain there to be transported by the current. These devices have been equipped to emit a sonic signal so that they may be followed by a surface ship. The deep water transport measured with these devices off Bermuda in the late 1950's and in 1960 was surprisingly high. Up to 42 cm/sec or a little less than 1 knot at a depth of 4000 meters.

C) Parachute drogues - Another method of measuring subsurface currents is the parachute drogue which has received a good deal of attention in the last ten years. With this method, a surplus air parachute ranging from 30 to 100 feet in diameter is opened at some pre-determined depth. It is suspended by wire from a small low drag surface float. The float is equipped with a radar reflector and a flashing light and is followed by a surface ship. The depth of the drogue may be determined by a recording pressure transducer attached above the parachute. Knauss (1963) has estimated that currents can

be measured to within about 5-10% of the surface current with this method. The drag force equation is: (Knaus)

$1/2\ C p\ Av^2$

C = drag coefficient (approx = 1)
p = density of the sea water
A = the cross sectional area of the wire and the float
V = velocity

D) Each of these "Flow" methods is expensive because of the ship's time required and each depends upon accuracy of navigation (a difficult problem in itself), but if the tagged particle is followed for a long enough period, the mean velocity can be calculated quite accurately.

II. The Eulerian Methods (or methods where the flow past a fixed point is measured) The greatest difficulty with this method is the non-existence of a "fixed" point, unless the instrument sits solidly on the bottom. In spite of this drawback, literally dozens of Eulerian current meters have been built and used. A brief discussion of some of the problems associated with this method is presented in the next section.

A) The Thorndike Bottom Photographic Current Meter sits on the bottom. It operates on the principle of a force deflecting a pendulum. Three flat pendula about the size of a large postage stamp are suspended vertically 120° apart and are deflected by the current. A calibration relating the magnitude of the deflection with the speed of flow is used to read the records, which appear on 35 mm photographic film. The film also shows a compass and a pointer which indicates direction of flow. The meter is mounted in a 900-lb. tripod which sets on the ocean bottom at any depth. The camera is started when a trigger weight reaches the bottom. Three secondary pendula indicate whether the tripod and meter are level on the bottom. A recent calibration of this meter indicates that it is capable of measuring currents down to about 2 cm/sec. Although restricted to measurements within a few feet of the bottom, this has proven to be a successful instrument. Future plans include the attachment to the tripod of a bottom camera which will photograph ripple marks on the bottom.

B) The Roberts Meter is a propeller-type meter where the number of revolutions of the propeller in a given time indicates velocity. It may either be suspended from a ship or hung below an anchored buoy on a conducting cable. An electric signal is transmitted to a surface recording device. The main disadvantage of this meter is that it readily responds to vertical motion and, therefore, is affected by any motion of the supporting platform.

C) Savonius Rotor Current Meters is perhaps the most popular type of Eulerian current meter in present use.

1) There are basically two types: an electrically operated deck recording meter which is lowered from the deck of a ship; a photographic recording long-period meter which is generally hung in arrays on anchored buoys.

2) The velocity sensor is a Savonius rotor which is magnetically coupled through a heavy-walled pressure case.

3) This type of meter has the advantage over the Roberts meter in that the Savonius rotor is relatively insensitive to vertical motion.

4) A recent series of tow tank experiments indicate that the Savonius rotor is sensitive from about 5 kts. down to about .1 kt. (5 cm/sec) but that its performance is dependent on its orientation relative to the axis of flow. Separate calibration curves must be prepared for the several possible orientations and the meter must be equipped to record its own orientation.

a: In the case of the photographic recording long-period current meter, this is accomplished by placing a compass and tilt indicator within the field of the camera.

b: For the electrically-operated deck recording meter, this problem has been minimized by hanging as much as 350 lbs. of lead ballast below the meter to keep the suspension wire vertical from the winch to the meter, and thereby reducing the number of necessary calibration curves.

III. Types of errors in direct current measurements. The problems of working at sea with the current meters are numerous and difficult to anticipate; any listing of sources of error is an over-simplification. The following is a summary of Paquette's (1963) excellent discussion of the problems.

A) Errors without motion of the platform

1) The near surface flow may be distorted by the platform; these effects may extend a ship's length downstream and are of particular importance where the velocities are small.

2) The magnetic compass in the current meter may deviation due to the iron on the platform. One hundred degree errors have been observed 70 ft. below a 275 ft. steel ship.

3) The elasticity of the system may cause a distorted response which can lead to a time lag in recording current fluctuation.

4) The wire angle or wire deflection leads to uncertainties in depth determination.

B) Dynamic Errors

1) Effects of slow or random motions of the platform and may seriously affect both velocity and direction measurements. These errors are expected to be smaller when the current meter is suspended below a buoy instead of hung from a ship.

2) Errors due to elasticity of long suspensions and the elasticity and slack in movings.

3) Errors due to dynamic failure of the meter itself. These prevent the meter from registering rapid transient velocities.

4) Errors due to pendula, elastic-cord and rotary types of oscillation which are due to the rolling and heaving of the platform.

5) Errors due to the vertical motion of the current meter. These are likely to be much greater in the case of the propeller-type meter than with the meter which uses a Savonius rotor.

References on Temperature and Depth Measurements with Deep-Sea Reversing Thermometers

Böhnecke, G. (1932) Die temperaturmessung an der oberfläche und in der tiefsee, METEOR Rep. 4, 1 (199-244)

Bumpus, D. F., (1948) The mechanics of correcting reversing thermometers and making depth determinations, WHOI

Culbertson, Margaret F., (1955) An oceanographic slide rule for computing temperature, depth and salinity: Transactions, American Geophysical Union, Vol. 38, No. 3 (473-480)

Defant, A. (1961) Physical Oceanography, Vol. I, Pergamon Press, P. 35

Dietrich, G. (1950) Uber Systematische Fehler in den beobachteten Wasserund Luft temperatures auf dem meere und ihre Answistung, Deutsche Hydrographische Zeitschrift, Band 3, Heft 5/6, (314-324)

Dietrich, G. (1963) General Oceanography, An Introduction, Interscience Publishers, John Wiley and Sons, p. 127

Ennis, C.C., (1944) Note on the computation of density of sea water and on corrections for deep-sea reversing thermometers: Observations and Results in Physical Oceanography, Scientific Results of CARNEGIE Cruise VII, Carnegie Inst. of Washington, Pub. 545, (23-46)

Feruglio, G. (1912) Il termometro a roveseiamento Richter e ta vole per la sua correzione, Memoria VII, (Reale) Comitato talassografico Italiano, Venizia.

Folsom, T.R., F. D. Jennings and R. A. Schwartzlose (1959) Effect of pressure upon the "protected" oceanographic reversing thermometer: Deep-Sea Research, Vol. 5 (306-309)

Folsom, T.R., R.A. Schwartzlose and F. D. Jennings (1962) Scale errors on oceanographic mercurial thermometers: Deep-Sea Research, Vol. 9 (219-226)

Geissler, H. (1931) Tiefenmessing mit ungeschutzden thermometern: Annalen der Hydrographie, Vol. 59 (433-438)

Gerard, W. G. (1933) Reversing thermometers: Hydrographic Review, Vol. 10, 1 (124-225)

Hamaker, H. C. (1941) The oceanographic instruments and the accuracy of the temperature observations, No. 4 Some notes on the discussion of errors, No. 5 The protected reversing thermometer and the temperature determinations, No. 6 The unprotected reversing thermometers and the determination of the depth of observation, No. 7 Comparison of the different sources of error and their influence on the interpretation of our temperature observations: Scientific Results of the Snellius Expedition, Vol. II, Part 1, Cahp. 1, (10-44)

Hansen, W. (1934) Bemerkungen zu den Korrektionsformein fur das Tiefsee Unkippthermometer: Annalen der Hydrographie, 1934, (145-147)

Herdman, H.F.P. and L.H. Pemberton (1958) The reliability of deep-sea reversing thermometers: DISCOVERY Reports, Vol. XXIX, (229-244)

Hidaka, K. (1932) Ueber eine neue korrektionsformula zur umkippthermomet erzblesung. Memoirs of the Imperial Marine Observatory, Kobe, Japan, 5, (11-13)

Houtman, T.J. (1961) A note on errors in depth calculated from unprotected deep-sea reversing thermometer readings: New Zeland J. of Geol. and Geophys., 4, 9 (415-423)

Keyte, F.K. (1964) On the formulas for correcting reversing thermometers, WHOI 64-29

Lafond, E. (1951) Processing oceanographic data, Hydrographic Office Publication 614, Washington, D.C. (29-39)

Nordstrom, S.G. and T.R. Folsom (1960) Suggestion for eliminating pressure effects on reversing thermometers: Deep-Sea Research Vol. 6, p. 169

Pollak, M.J. (1950) Notes on determining the depths of sampling in serial oceanographic observations: J. Mar. Res., Vol. IX, No. 1 (17-20)

Reid, R.O. (1949) A theoretical study of the slope of the hydrographic wire: Scripps Inst. of Oceanography (unpublished manuscript)

Reid, R.O. (1958) Influence of some errors in the equation of state or in observations on geostrophic currents: Physical and Chemical Properties of Sea Water, National Acad. of Science - National Research Council Publication 600, (10-31)

Richter, N., Über die Funktion der Umkippthermometer sowie die Beseitigung von Störungen: Richter and Wiese Co., Berlin, Germany

Ruppin, E., (1906) Umkippthermometer als Tiefenmesser: Wissenschaftliche Meeresuntersuchungen N.F., Vol. 9, 5, Kiel Section

Schumacher, A., (1923) Neue Rilfstafeln fur die Umkippthermometer nach Richter und Beitrage zur thermometrischen Tiefenmessung: Annalen der Hydrographie, 1923 (273-280)

Soule, F.M. (1944) Note on the practical correction of deep-sea reversing thermometers and the determination of the depth of reversal from protected and unprotected thermometers: Observations and Results in Physical Oceanography, Scientific Results of CARNEGIE Cruise VII, Carnegie Institution of Washington Publ. 545 (19-22)

Sverdrup, H. U., M.W. Johnson and R.H. Fleming (1942) The Oceans, Their Physics, Chemistry and General Biology, Prentice Hall p. 350

Sverdrup, H.U. (1944) Observations and Results in Physical Oceanography, Scientific Results of CARNEGIE Cruise VII, Carnegie Institution of Washington Publ. 545

1. Sub-surface temperatures (5-10)
2. Thermometric determination (11-14)
3. Determination of depths at which temperatures were measured and water samples collected, (15-18)
4. Depth to bottom at CARNEGIE stations, (47-50)

Sverdrup, H.U. (1947) Note on the correction of reversing thermometers: Journ. of Marine Research, VI, 2 (136-138)

Whitney, G.G. Jr. (1960) Procedure for and comments on ice point tests of deep-sea reversing thermometers, WHOI Contrib. No. 1064

Whitney, G.G. Jr., (1952) Notes on malfunctional behavior and its correction in deep-sea reversing thermometers WHOI Ref. No. 52-29

Whitney, G.G. Jr. (1952) Comments on the determination of the pressure factor "Q" of unprotected reversing thermometers: WHOI Ref. 52-30

Whitney, G.G. Jr. (1955) A note on the non-linear behavior of unprotected reversing thermometers WHOI Ref. 55-62

Whitney, G.G. Jr. (1957) Factors affecting the accuracy of thermometric depth determinations: Journal du Conseil International pour l'Exploration de la Mer, XXII, 2 (167-173)

Wüst, G., (1932) Thermometric measurement of depth: Hydrographic Review, 10, 2 (28-49

Wüst, G., (1933) Thermometrische Tiefenmessung: Wissenschaftliche Ergebnisse der Deutschen Atlantischen Expedition auf dem Forschungs- und Vermessungschiff METEOR, 1925-27, Pub. by A. Defant, Section B of Vol. 14, Part 1, Berlin and Leipzig.

Wüst, G. (1961) Lectures on Problems of Physical Oceanography, Part A, Chap. 1-VIII, Lamont Geological Observatory (unpublished manuscript)

References on the Direct Measurement of Ocean Currents

Böhnecke, G. (1955) The principles of measuring currents, Association d'Oceanographie Physique: Publication Scientifique No. 14, Geofysik Institut, Bergen

California, State of, State Water Pollution Control Board, Publication No. 19, Investigation of current measurement in estuarine and coastal waters. This contains the most complete bibliography on all aspects of direct current measurements previous to 1959.

Buoy Technology, An aspect of observational data acquisition in oceanography and meteorology, Marine Technology Society, 1964, Various papers (Also, see supplement to this publication)

Dietrich, G. (1963) General Oceanography - An Introduction Interscience Publishers, Part III, Oceanographic Instruments and Observational Methods Part C Current Measurements

Gaul, R.D., J.M. Snodgrass and D.J. Cretzler (1963) Some dynamical properties of the Savonius rotor current meter: Marine Sciences Instrumentation, 2 (115-126 Plenum Press, New York

Gerard, R., (1963) Direct current measurements and hydrographic observations in the Caribbean-Antillean Region: Lamont Geological Observatory Tech. Rep. CU-5-63 (unpublished manuscript)

HyTech Products, The Bissett-Berman Corp., a report on the Model 9006 in situ salinity-temperature-depth monitoring and recording system, Tech. Rep. 102

Knauss, J.A. (1963) Drogues and neutral-buoyant floats, Chap. 14, The Sea, Vol. 2 Interscience Publishers, John Wiley and Sons, New York

Paquette, R.G. (1963) Practical problems in the direct measurement of ocean currents: Marine Science Instrumentation, Vol. 2 (135-146) Plenum Press, New York

Parker, Charles E., (1963) The radio drogue system: WHOI Ref. 63-12

Richardson, W.S., P.B. Stimson and C.H. Wilkins, (1963) Current Measurements from moored buoys: Deep-Sea Res. 10, 4 (369-388)

Sexton, R. (1964) Some tow tank calibrations of the Savonius rotor: Lamont Geological Observatory Tech. Rep. CU-11-64 (unpublished manuscript)

Thorndike, E.M. (1962) A suspended drop current meter: Deep-Sea Res. 10 (263-267)

Volkmann, G.H. (1963) Deep current measurements using neutrally buoyant floats The Sea, Vol. 2 Interscience Publishers, John Wiley and Sons, New York

Von Arx, William S. (1962) An introduction to physical oceanography, Addison-Wesley Pub. Co., Reading Mass. Chapters 8 and 9

Processing Oceanographic Data, Core Method and Geostrophic Method

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Introduction

This section of the workshop is mainly concerned with the geostrophic method, as used in oceanography. This method is the first step towards a quantitative understanding of an oceanic region. It is the calculation of flow, its direction and magnitude, from the internal field of mass. Before a geostrophic study can be performed, the general circulation must be known in a qualitative way. This is necessary for a better choice of which station pairs to use, that is, station pairs or profiles which are nearly perpendicular to the main flow. A qualitative understanding of an area is vital in planning future cruises. The best method to use in gaining this qualitative understanding is the core method, as developed by Georg Wüst. Previous to both the core and geostrophic methods, the data must be inspected, suspicious data removed and if used in the interpretation, it must be marked as being doubtful. A brief outline covering the processing of oceanographic data and the core method precedes a more complete outline of the geostrophic method. The first two parts are followed by a bibliography, and the bibliography for the geostrophic part is found at the end of this section.

Processing Oceanographic Data

I. Aims of processing the data

A) After the mass of temperature data is corrected for scale imperfections and the conductivity values converted to salinity, the density is calculated. With these three parameters, the quality of the data is inspected. The reliable and suspicious data are separated. It would usually be found that the ratio of reliable to suspicious data varies from ship to ship.

B) Data is processed and put into a standard format for easy handling in the interpretation. The temperature, salinity and density values are recorded at observed levels and standard levels. As a standard part of the processing, sound velocity and dynamic height anomaly can be calculated. For each station, the particulars as date, time, sea and swell conditions, wind, etc. are included. In this standard form, the data is published.

II. Accuracy check

A) The basic assumption in checking data accuracy is the belief that density increases smoothly with depth and that curves relating temperature, salinity, density and depth are also "smooth". The final decision of either calling data

reliable or suspicious is reached by reasoning which is slightly subjective, and so it is best to have someone experienced in oceanography to perform the accuracy check. Recent work with continuous temperature and salinity recorders shows that the temperature vs. depth curves as well as the salinity vs. depth curve are not always smooth but in certain regions show minor maxima and minima of only a few meters thick. The chances of these secondary variations combining to give a smooth vertical density gradient with no "wiggles" are slight. It is recommended that data with obvious inconsistancies be discarded, but the data which seems to be only slightly incompatible be marked as doubtful and used as such. The accuracy check is a very important step in the processing of the data and great care by an experienced oceanographer must be taken that new important findings are not discarded as unreliable data. It is also advisable that the person who gathered the data perform this check since he knows the behavior of the instruments used and the conditions under which the data was taken.

B) The temperature vs. salinity diagram is in general use for the accuracy check of the data. Each area would have a characteristic T-S curve which may vary with the season. Data taken in an area must fall within an envelope which surrounds this curve. If the area is in nearly steady state conditions, the envelope would be as wide as the accuracy to which temperature and salinity can be measured; if the area has a real variability in its parameters, the envelope would be wider. If the new data falls in the envelope and leads to a fairly smooth T-S curve, it is taken as reliable; if it falls in the envelope but does not give a smooth curve, it is marked as doubtful; if it falls outside of the envelope, it is usually bad and is discarded. The density lines which appear as diagonals are superimposed on the temperature-salinity grid. These lines help distinguish any improbable density gradients.

C) A method developed by Hans Klein is used by the National Oceanographic Data Center as a quality control. In this method, two curves are drawn: σ_t - Z and the σ_t - S o/oo. The T-S envelopes can be converted to σ_t - S coordinates. This method may be superior to the standard T-S diagram since the check is based directly on the σ_t - Z curve, which should be the smooth and so the easiest to construct. Temperature and salinity can both be incorrect, but give a density which "fits" the σ_t - Z curve, in the case the σ_t - S curve would not be smooth. The probability of this is slight. A recent study (Morey, 1963) has demonstrated that 95% of errors can be attributed to salinity, so it is more likely that both curves will be distorted. This method can be used to fill in points where only temperature or salinity is known and for retrieval of "bad" points. It is also an excellent way to perform a graphic interpolation to standard levels, or as a check on machine-interpolated values.

D) When the σ_t - S o/oo; σ_t - Z diagrams are used in conjunction with the standard T - S diagram, inconsistancies are revealed which were not apparent on one independently.

The Core Method - (Kernschichtmethode)

I. Basic Principles

A) As the meteorologist recognizes various source areas for the production of types of air, the oceanographer also identifies regions as producing a certain type of sea water. These source areas are regions where the fluid (air or sea water) obtains a characteristic trait; it flows from this area and gradually loses this trait by mixing or through the action of some external influences. An example of this in meteorology is the formation of a continental polar air mass in northern Canada with its characteristic low temperature and low water vapor content; as this mass of air moves south, it gradually warms and becomes wetter due to contact with the land and perhaps some mixing with the warm tropical air masses to the south. A similar example in oceanography would be the formation of the Antarctic bottom water originating in the continental shelf of the Antarctic continent; it sinks to the ocean bottom and spreads north. It is characterized by low potential temperature and higher salinity than the surrounding water. It gradually loses its identity through mixing with the North Atlantic Deep water and possibly through warming due to heat flow from the earth.

B) The water produced at a source area would have an extrema in one or more parameters. It will spread from this area at a depth which is determined by its density and the vertical density column of the surrounding waters. The layer in which this characteristic water flows is called the core layer, which can be found by inspecting the vertical distribution of parameters for extrema; this layer is not confined to remain at constant depth. An assumption of steady state in the ocean's general circulation must be made in this type of study, as must be also done in many other classical approaches. Wüst (in his core studies in the Atlantic Ocean) has shown that this assumption is valid and that in a core study of an oceanic region, data may be treated synoptically.

C) The following is a table giving the parameters used to identify the main core layers in the cold water sphere of the Atlantic Ocean.

Core Layer	Source	Parameter
Arctic and Antarctic Bottom Water	Polar regions	Potential bottom temperature minimum
Middle and Lower North North Atlantic Deep Water	Area near Greenland	Intermediate maximum of oxygen
Upper North Atlantic Deep Water	Mediterranean Sea	Intermediate maximum of salinity
Subantarctic Intermediate Water	Antarctic Polar Front	Intermediate minimum of salinity

II. The mechanics of a core study

A) All the oceanographic data in the area of interest is collected and checked for reliability (See Part I of this section of the Workshop).

B) Each extrema which is common to most stations is plotted on an areal map where isolines can be drawn and the main axis of spreading located. Vertical profiles, parallel and transverse to this main axis are prepared. There is usually more than one core layer, and an areal map must be prepared for each. Wüst has found in his latest study, "Stratification and Circulation in the Antillean-Caribbean Basins" that the data for the oxygen maximum core layer in the Caribbean cannot be treated synoptically and must be broken into periods, but the other major core layers in the Caribbean Sea were found to be fairly constant in thickness and depth of occurence for the long period of data collecting in the area.

C) The vertical and horizontal diagrams are not confined to flat walls or surfaces of constant depth, but follow the surface of the core layer itself; thus, each diagram represents the actual spreading of the sea water. Wüst has found that the core layers in the Caribbean have a general slope from south to north and by preparing areal maps which follow the sloping core, he was able to define a very clear main axis of spreading which was later supported by geostrophic calculations.

D) If the core layers can be identified as having formed at a particular source area, points along the main axis of spreading can be plotted in a T-S diagram and the percent of the original water remaining can be found as a function of the distance from the source region.

III. The above is just a brief introduction to the core method. For a complete treatment, the reader is referred to the works of Georg Wüst. (See references).

BIBLIOGRAPHY

Defant, A. (1961) Physical Oceanography, Vol. 1, Pergamon Press

Klein, H. (1956) Graphical presentation of a hydrographic station and dynamic computations. SID-MLR Form 4, 5. University of California, Scripps Institution of Oceanography, La Jolla, California

LaFond, E. C. (1951) Processing Oceanographic Data, U. S. Navy Hydrographic Office Publication No. 614, U. S. Navy Hydrographic Office, Washington, D. C.

Montgomery, R. and Wooster, W. (1954) Thermosteric anomaly and the analysis of serial oceanographic data. Deep-Sea Research, Vol. 2, No. 1, Oct., 1954

Morey, C. (1963) A study of the water mass of the Tongue of the Ocean, University of Rhode Island, Master of Science Thesis (1963). Unpublished manuscript

Sverdrup, H. U., Martin W. Johnson and Richard H. Fleming (1942) The Oceans, Their Physics, Chemistry and General Biology, Prentice-Hall, Inc.

U. S. Navy Oceanographic Office (1960) A comparative study of three interpolation techniques used in processing oceanographic station data. 10M 28-60. Unpublished manuscript.

Winterfield, T. and Carl F. Morey (1963) The salinity-density relationship as a tool for the interpolation and quality control of oceanographic station data, A Progress Report by the National Oceanographic Data Center

Wooster, W. S. and Taft, B. A. (1958) On the reliability of field measurements of temperature and salinity in the ocean. J. Mar. Res. 17, 552-566

Wüst, G. (1935) Die Stratosphäre: METEOR Exped. 1925-27; Wiss. Erg., Bd 6 Teil 1, Lief 2, Berlin

Wüst, G. (1936) Das Bodenwasser und die Gleiderung der Atlantischen Tiefsee, METEOR Exped., Band 6, Teil 1

Wüst, G., (1961) On the vertical circulation of the Mediterranean Sea, Jour. of Geophys. Res., Vol. 66, No. 10

Wüst, G., (1964) Stratification and Circulation in the Antillean-Caribbean Basins, Part I, VEMA Research Series No. II, Columbia University Press

The Geostrophic Method in Oceanography

I. The Theory

A) Basic equation

1) In 1900 Bjerknes developed a method of determining the flow of a fluid from its internal field of mass. Helland-Hansen (1903) modified the method for easy use in oceanography. He arrived at the form

$$(V_0 - V_1) = \frac{N}{fL}$$

where: $(V_0 - V_1)$ is the velocity difference between the reference level of velocity V_0 and the level of interest V_1; f is the Coriolis parameter $(2\,\omega \sin\phi)$; L is the distance between the pair of stations: N is the number of solenoids enclosed by the loop of integration. Solenoids are "tubes" formed by intersection of isobaric with isosteric surfaces. In the CGS system $N = (D_A - D_B) \times 10^5$ where D_A and D_B are the anomaly of dynamic height in dynamic meters at stations A and B respectively. The units of fL, V_0 and V_1 are cm/ sec.

2) The geostrophic relation can be derived from the Eulerian equations of motion, in Cartesian coordinates:

$$fV = \alpha \frac{\partial p}{\partial x} = \left.\frac{\partial D}{\partial x}\right)_p$$

$$fU = -\alpha \frac{\partial p}{\partial y} = -\left.\frac{\partial D}{\partial y}\right)_p$$

in Spherical coordinates:

$$\frac{V_\lambda^2}{r}\tan\phi + fV_\lambda = -\frac{\alpha}{r}\frac{\partial p}{\partial \phi}$$

$$\frac{V_\phi V_\lambda}{r}\tan\phi - fV_\phi = -\frac{\alpha}{r\cos\phi}\frac{\partial p}{\partial \lambda}$$

B) Assumptions and their validity

1) The assumptions of geostrophic balance are more explicit in the equation of motion treatment than in Bjerknes approach. The terms which are ignored are the following:

a: Lateral and vertical stresses, $N_H \dfrac{\partial^2 V}{\partial H^2}$ and $\dfrac{\partial \tau_{zH}}{\partial z}$ respectively.

b: Time derivative of the velocity

c: Advection terms

d: Vertical velocity

2: Validity of the assumptions can be studied by converting the equations of motion to dimensionless form (See Dr. Ichiye's section) or by simply calculating typical oceanic valaes for the various terms, as done below. The Coriolis' term is about 10^{-3} - 10^{-4} in mid-latitudes and 10^{-4} - 10^{-5} near the Equator.

a: Stresses - Due to the external force of the wind, the vertical stress term is important in the upper 200 meters; it is also important in the lower few hundred meters because of the frictional influence of the bottom. The order of magnitude in these two layers can be as high as 10^{-5} cm/sec^2 and probably 10^{-4} cm/sec^2 in the immediate vicinity of the two boundaries. Between these two frictional layers, the vertical stresses are rarely above 10^{-6} cm/sec^2. Lateral stresses ($N_H \frac{\partial^2 V}{\partial H^2}$ where N_H is the horizontal eddy viscosity) can be as high as 10^{-4} cm/sec^2 in the regions of high shear of the boundary currents, but beyond these boundaries, the lateral stresses are usually less than 10^{-6} cm/sec^2. The effect of stress can be neglected when compared to the Coriolis' term at most areas except at the ocean's boundaries and near the Equator.

b: Time derivative: When a motion has a periodicity much less than that of the rotational period of the earth, the flow would not be geostrophic. It is when this period is of the same order or greater than that of the earth's period that balance is obtained. For motions with large periods, the acceleration term is much smaller than the Coriolis' term. As an example, the particle motion of an internal wave with a period the same as the earth's rotational period would be (6×10^{-10}) multiplied by the amplitude of the wave. Thus, a wave would have to have an amplitude of over a kilometer for the time derivative to be of importance.

c: Advection terms: These terms are of importance where velocity shear is high. Values in mid-ocean are of the order of 10^{-7} to 10^{-8} cm/sec^2, but in the Gulf Stream values can be as high as 10^{-3}cm/sec^2.

Thus, the advection terms can be neglected except at the boundaries.

d: Vertical velocity: It is assumed that $W \cos \phi << V \sin \phi$; this is true except at the Equator. It is also assumed that $W \frac{\partial U}{\partial z}$ and $W \frac{\partial V}{\partial z}$ can be neglected; this is valid in most places except where upwelling occurs and where large values of vertical shear is encountered as on the Equator.

C) Areas where the validity of the geostrophic relation is doubtful

1) The validity of the relation becomes more questionable as we approach the Equator. Knauss has shown balance to within 1/2 degree from the Equator. However, geostrophic balance cannot be expected this close to the Equator at all levels and all equatorial regions. Near the ocean's vertical and horizontal boundaries where frictional and advection terms can be large, the forces may not be in geostrophic balance in all directions; that is, the meridional forces may lead to geostrophic flow, while the latitudinal forces do not.

2) In weak baroclinic or barotropic fields where both pressure gradients and Coriolis' terms are small. This is true for two reasons: due to the relative sizes of the terms in the equation of motion and secondly, due to the limits of accuracy that horizontal density gradients can be measured.

D) The geostrophic component of the velocity field

The flow necessary for balance of Coriolis' term to the mean pressure gradient is the geostrophic component of velocity. Mean steady state conditions are geostrophic, but some transient effects can also be geostrophic, as Rossby waves which are essentially a moving wave of geostrophic flow. The Rossby number is a useful parameter for relating velocity and dimension of a feature to geostrophicity.

II. The calculation - The flow in the ocean is essentially geostrophic, but we measure the pressure gradients to sufficient accuracy to make use of this balance?

A) The ocean is divided into two parts: a standard component (35 o/oo, 0°C) and an anomalous component, since the standard ocean contains no solenoids, it is only the anomalous part which is of concern. The anomaly of dynamic height can be written as

$$D = \int_0^P \delta \, dP$$

where δ is the anomalous component of the in situ specific volume.

It is defined as

$$\delta = \alpha_{s,t,p} - \alpha_{35,0,p} = \delta(s,t,p)$$
$$\delta_{(s,t,p)} = \Delta_{s,t} + \delta_{s,p} + \delta_{t,p}$$
$$\Delta_{s,t} = \Delta_{s,T}(\sigma_t)$$

Since we have only point measurements of δ , we cannot carry out the integration exactly and an approximation is used; the Trapezoidal Rule is employed for this purpose.

B) Temperature and salinity at the observational points are interpolated to standard depths; the resultant δ are used in the dynamic calculations. There are various ways in which this interpolation can be carried out.

1) Graphical means give smooth curves, but this method is time consuming. The interpolation can be done by computer. The scheme most commonly used is the Lagrange interpolating curves. These are curves of the nth degree using n + 1 pivotal points. Three-point and four-point curves have been used. A modified 3-point method has been developed by Rattray. This method uses two 2nd degree curves and an average value is chosen. Other schemes can be used, such as linear interpolation (under special conditions) and various matrix and log methods. A combination of graphical and 3-point curve is suggested by the NODC.

2) The various schemes of interpolation can yield different standard depth values for the same stations; these are discrepancies rather than errors. Since the true values at these levels are not known, the hand or graphical method is chosen to be correct and the computer curve with the smallest discrepancy from this curve is said to be best. An alternate approach is to say no one curve is better than the next and standard depth values are best determined when discrepancies between curves are small.
If no reversals occur in T-Z and S-Z curves and the bottle spacing is sufficiently dense, linear interpolation can be used. In a Gulf Stream station (CRAWFORD 230) the discrepancy between a linear scheme and the graphical is only 2 dy cm in 4000 meters. If reversals occur, discrepancies would be higher, as seen when linear and the modified 3-point methods are compared for the Caribbean station CRAWFORD 341; here, there is a discrepancy of 3 dy cm in 3500 meters.

Lagrange methods for interpolations can go astray at reversals, and editing of the computer program is done

to prevent extreme values. A study carried out by the Hydrographic Office in 1960 compared three methods by leaving out observed points and interpolating for them using the neighboring points. The methods compared were a 3-point Lagrange, a log method and a hand drawn curve. Discrepancies in sigma-t calculated from the interpolated temperature and salinity were approximately 0.02 units in the upper 350 meters and zero below 350 meters. The error from the observed values were quite high, about 0.1. This method of study is not the best since it cuts down on the density of observations. In a recent report a 4-point Lagrange scheme is discussed. With heavy editing of the program, the computer was within 0.01 units of sigma-t (in 95% of the cases) when compared to a hand-drawn curve.

C) Uncertainty in our knowledge of the correct coefficients relating density to temperature, salinity and pressure may lead to large errors in in situ specific volume. The compressibility coefficient error is not important since we compare specific volumes on similar levels. Errors in the thermal expansion coefficient can be important since horizontal temperature gradients exist. The thermal expansion coefficient is known to a fair degree of accuracy in the pressure range of 1 to 1000 db; above this range its value is in question. This possible error is not of great concern since the horizontal temperature gradients are small below 1000 db (984 meters). Reid believes the error in dynamic height of the sea surface based on a deep reference level is approximately 5% of the true value. This is smaller than the errors due solely to inaccuracies of our measurement of temperature and salinity.
The tables of specific volume anomaly can be found in The Oceans and LaFond's H.O. Publication #614

D) Statistical study of the geostrophic method

1) The accumulative dynamic height anomaly has an associated uncertainty due to errors in measuring temperature and salinity. The errors in calculating the anomaly of specific volume can be expected to fall in a Gaussian ~~anomal~~ normal distribution

$$f(Z) = \frac{h}{\sqrt{\pi}} e^{-h^2 Z^2}$$

where f(Z) is the Gaussian curve and e is the precision constant. The larger the value of h, the steeper the Gaussian curve. This precision constant is a function of the error in δ measurement, the number of bottles between the pressure surface of interest and the reference layer, and the distance in the reference level. The probability of an error lying in the range -Z to Z is given by the integral

$$W_{-z,z} = \int_{-z}^{z} f(x)\,dx = \frac{2}{\sqrt{\pi}}\int_{0}^{hz} e^{-t^2}\,dt$$

This integral is tabulated in many texts on statistics and from it, the various confidence levels can be found.

2) Fomin has made a study using a temperature error of ± 0.02°C and a salinity error of ± 0.009 o/oo. The bottle spacing used was 25 meter intervals from 0-200 m, 100 meter interval from 200-1000 m, and 500 meter interval from depths below 1000 m. I have completed a statistical study using a 90% confidence level error of 1.10 x 10^{-5} in the anomaly of specific volume.* Like Fomin, I have assumed no error in the depth of the temperature and salinity values. I also used more standard levels: 50 meter intervals from 0-700 m, 100 meter intervals from 700-2000 m, and 250 meter intervals below 2000 m. The following table gives the probable error of the anomaly of dynamic height (in dynamic millimeters) of the surface pressure for a reference layer at various depths.

Table I

Depth of Reference Layer	Probable Error (dy mm.) after Fomin	Probable Error (dy mm.) after Gordon
600	2.1	0.8
700	2.4	0.8
800	2.5	0.9
900	2.7	1.0
1000	2.9	1.1
1500	4.2	1.5
2000	6.4	1.8

3) The next table gives the 90% confidence level of the dynamic height anomaly (column A) and of the $(\Delta c)(fL)$ value (column B). The values in column B must be divided by the fL value for the station pair to convert to the 90% confidence level in velocity. Both columns assume a reference layer at 1000 meters.

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* This study is to appear in VEMA research series No. III which should be published in 1965.

Table II (after Gordon)

Depth (m)	Column A 90 % confidence level for dynamic height (dy mm.)	Column B 90% confidence level for $(\Delta c)(fL)$ in $(cm/sec)^2$
0	2.73	3.86 $\times 10^2$
50	2.66	3.76
100	2.61	3.69
150	2.56	3.62
200	2.49	3.52
250	2.44	3.44
300	2.37	3.34
350	2.32	3.27
400	2.24	3.17
450	2.17	3.08
500	2.10	2.98
550	2.02	2.86
600	1.95	2.76
650	1.87	2.66
700	1.76	2.49
800	1.34	1.90
900	0.78	1.10
1000	0.0	0.0
1100	0.78	1.10
1200	1.34	1.90
1300	1.76	2.49
1400	2.07	2.92
1500	2.34	3.29
1600	2.58	3.66
1700	2.80	3.95
1800	3.00	4.25
1900	3.22	4.56
2000	3.39	4.78
2250	4.10	5.81
2500	5.03	7.10
2750	5.65	8.00
3000	6.30	8.88
3250	6.89	9.36
3500	7.42	10.48
3750	7.90	11.15
4000	8.37	11.80

4) The accuracy to which the geostrophic velocity can be found is dependant on the fL value. It is best to have the separation of the hydro stations as far apart as the scale of the current features would allow. If the stations are too far apart, permanent eddies may be missed and if too close together, the error would be large. The relative size of the error to the absolute velocity depends upon the magnitude of the velocities

present; thus, the best results would be expected in regions of strongly sloping pressure surfaces. Where the flow is weak, the errors may be more than 100% of the true velocities making the direction of the flow uncertain.

III. The reference Level - To find absolute velocities, we must know the geostrophic flow at some level; the flow at this level can be of a finite value or zero.

A) Reference levels found by indirect means are levels or layers of no motion. They are levels of no horizontal motion in any direction, or just in the direction perpendicular to the station profile. The following are the more important methods that have been used.

1) The oxygen minimum which exists in intermediate depths in some parts of the world ocean is thought by some to indicate a layer of little horizontal motion. Dietrich has used this method in geostrophic calculations of the Gulf Stream. Seiwell does not believe that this minimum is a zero reference layer and has shown (1937) that this is not a necessary condition for its existence.

2) Parr (1938) suggests that the zero level can be found by plotting $\frac{\Delta Z}{\Delta \sigma_t}$ against $\overline{\sigma_t}$ for a number of hydro stations in the same area. The $\overline{\sigma_t}$ which has similar $\frac{\Delta Z}{\Delta \sigma_t}$ for all the stations would be the zero reference level. He argues that a layer with no horizontal motion will have an undistorted isopycnal layer (where an isopycnal layer is a layer formed by two different surfaces of equal σ_t). The density difference between two stations for depths below 1500 m is small and so if the reference layer is below this depth, all the $\frac{\Delta Z}{\Delta \sigma_t}$ would be nearly equal. Horizontal density gradients increase with decreasing depth so unless the zero layer is shallow, we may not have the accuracy to measure differences in $\frac{\Delta Z}{\Delta \sigma_t}$.

3) The salt diffusion equation for a steady state condition is given by

$$A_H\left[\frac{\partial^2 s}{\partial x^2}+\frac{\partial^2 s}{\partial y^2}\right]+A_Z\frac{\partial^2 s}{\partial z^2}=\frac{u}{\rho}\frac{\partial s}{\partial x}+\frac{v}{\rho}\frac{\partial s}{\partial y}+\frac{w}{\rho}\frac{\partial s}{\partial z}$$

At the zero reference level, Hidaka broke this expression into two parts -

$$\frac{\partial^2 S}{\partial x^2} + \frac{\partial^2 S}{\partial y^2} = 0 \quad \text{for predominates of horizontal diffusion}$$

$$\frac{\partial^2 S}{\partial z^2} = 0 \quad \text{for vertical diffusion}$$

He applied these two equations to the Atlantic and found the first gave no reference level, and the second gave one ranging from 900 m to 1400 m. An objection to this method is that we did not know if it is the Laplacian of the salt distribution or the coefficient of eddy diffusion which is equal to zero when no motion is present. What does turbulent diffusion mean in the absence of motion ? This objection does not hold with $\frac{\partial^2 S}{\partial z^2}$ equation since $\frac{\partial u}{\partial z}$ can be present, and so $A_Z \neq 0$.

4) The most reliable method for determination of the zero reference layer is Defant's method. It is basically the belief that the layer of zero horizontal velocity also is a layer with weak vertical velocity gradient. A plot is made of $D_A - D_B$ vs. depth, the zero reference layer is placed where $D_A - D_B$ does not change with depth. The velocity above this reference layer must have the opposite sign of that below; this is a necessary restriction to apply since a zero $\frac{\partial v}{\partial z}$ with the same direction of flow above and below would more likely be a layer of maximum flow.

5) The continuity equation can be used to find a reference layer: as much water must flow into a water column as flows out. Thus, by moving a reference level vertically until this condition is fulfilled, the zero level can be reached. The objection to this is that we cannot calculate deep currents to sufficient accuracy for determination of a reliable volume transport. The method was originally suggested by Hidaka in 1940.

6) Stommel has applied the continuity equation and the conservation of potential vorticity to find the layer of no meridional motion

B) Direct current measurements can be used as a reference. A long record of the instantaneous current is needed so harmonic analysis will be able to separate the fluctuations from the mean flow. It is best to place the current meter in a depth of little shear where fluctuation due to turbulence would be a minimum.

BIBLIOGRAPHY

Arthur, Robert (1960) A review of the calculation of ocean currents at the Equator. Deep-Sea Res., Vol. 6

Bjerknes, V. (1900) Das dynamische prinzip der zirkulationsbewegung in der atmosphäre. Meteorol. Z.

BJerknes, V. and J. W. Sandström (1910-1913) Dynamische meteorologie und hydrographie, 2 vol. Vieweg Braunschweig. English edition by Carnegie Institution No. 81, Washington 1910-1013.

Charney J. G. (1947b) On the scale of atmospheric motions, Geofys. Publikasjoner, Norske Videnskaps-Akad. Oslo, 17 (2).

Defant, A., (1961) Physical Oceanography, Vol. I, Pergamon Press

Defant, A. (1941) Die absolute topographie des physikalischen meeresniveaus und der druckflächen, sowie die wasserbewegungen im Atlantischen Ozean, Deutsche Atlantische Exped. "METEOR" 1925-27, Vol. 6, Part 2, No. 5

Dietrich, G., (1936) Aufbau und bewegung von Golfstrom und Agulhasstrom. Naturwissenschaften, No. 15.

Dietrich, G., (1963) General Oceanography, an Introduction. Interscience Publishers.

Dobrovol'skii, A. D. (1949) Contribution to the problem of the location of the zero surface for dynamic computation in the North Pacific. Tr. Inst. Okeanol. Akad. Nauk. USSR, V. 4

Eckart, C. (1958) The equation of state of water and sea water at low temperatures and pressures. Amer. J. Sci. 256, (225-240)

Ekman, V. W. (1905) On the influence of the earth's rotation on ocean currents. Arkiv Matem., Astr. Fysik, Stockholm, 2, 11

Farrell, J. (1964) A modified four-point Lagrangian interpolation method for oceanographic station data, TM No. 312, U. S. Naval Underwater Ordnance Station, Newport, R. I.

Fofonoff, N. P. (1962) Dynamics of ocean currents, Sect. III, Vol. 1, The Sea Interscience Publishers

Fomin, L. M. (1964) The dynamic method in oceanography, Elsevier Publishing Company

Godske, C. L., T. Bergeron, J. Bjerknes and R. C. Bundgaard (1957) Dynamic Meteorology and Weather Forecasting, Waverly Press, Inc.

Gordon, A. L., Quantitative study of the dynamics of the Caribbean Sea, VEMA Research Series No. III, Columbia University Press (to be published)

Greenspan, H. P. (1963) A note concerning topography and inertial currents, J. Mar. Res, 21, 3

Groen, P. (1948) Methods for estimating dynamic slopes and currents in shallow water, J. Mar. Res. 7, 3

Haiami, S., Kawai, H. and Ouchi, M. (1955) On the theorem of Helland-Hansen and Ekman and some of its applications, Rec. of Oceanogr. Works in Japan, V. 2, No. 2

Hidaka, K. (1940) Absolute evaluation of ocean currents in dynamical calculations, Proc. Imp. Acad-, Tokyo, V. 16

Hidaka, K. (1949) Depth of motionless layer as inferred from the distribution of salinity in the ocean, Trans. Am. Geophys. Union, Vol. 30, No. 3

Hidaka, K., (1950) An attempt to determine the motionless layers in the ocean. Oceanogr. Mag., V. 2, No. 1.

Hydrographic Office (1960) A comparative study of three interpolation techniques used in processing oceanographic station data.

Jakhelln, A., (1936) The water transport of the gradient currents. Geofys. Publikajoner, V. 11, No. 11

Jeffreys, H., (1926) On the dynamics of geostrophic winds, Quart. J. Roy. Meteorol. Soc., 52, (85-104)

Knauss, J. (1960) Measurements of the Cromwell Current, Deep-Sea Res., Vol. 6, (265-285

Leipper, D. (1959) The classical hydrographic procedure and its relation to the physical and chemical properties of sea water, Publication #600, Nat. Acad. Science and Nat. Res. Council.

Mamayev, O. L. (1955) Methods of determining the zero dynamic surface in the World Ocean., Vestn. Mosk. Univ., No. 10

McGill, D. (1961) The calculation of specific volume anomalies and interpolation for dynamic heights with the Recomp II computer, WHOI Tech Rep. No. 61-21

Montgomery, R. B. and W. Wooster (1954) Thermosteric anomaly and analysis of serial oceanographic data, Deep-Sea Res., Vol. 2 (68-70)

Neumann, G., (1942) Die absolute topographie des physikalischen meeresniveaus und oberflächenströmungen des schwarzen meeres. Ann. Hydrograph. Marit. Meteorol., V. 70, No. 9

Parr, A., (1938) Analysis of current profiles by a study of pycnomeric distortion and identifying properties, J. Mar. Res., No. 4

Phillips, N. (1963) Geostrophic motion, Reviews of Geophysics, Vol. 1, No. 2 (123-176)

Rattray, M., (1962) Interpolation errors and oceanographic sampling, Deep-Sea Res., Vol. 9, (25-37)

Reid, J. L. (1948) Influence of some errors in the equation of state or in observations on geostrophic currents. Conference on Physical and Chemical Properties of Sea Water, Nat. Acad. Science - Nat. Res. Council, Publication 600, (10-29)

Seiwell, H. R. (1939) The effects of short-period variations in dynamic oceanography, Pap. Phys. Ocean and Meteor., Vol. VII, No. 3

Shtokman, V. B., (1953a) Some problems concerning the dynamics of sea currents Izv. Akad. Nauk. USSR, Ser Geofiz. No. 1

Somov, M. M., (1937) Computation of current velocities between hydrological stations of various depths by the dynamic method. Meteorol. i Gidrol. No. 8

Stommel, H. (1955) Lateral eddy viscosity in the Gulf Stream system, Deep-Sea Res. No. 3, (88-90)

Stommel, H., (1956) On the determinations of the depth of no meridional motion, Deep-Sea Res. No. 3, 273-278

Stommel, H. (1957) A survey of ocean current theory, Deep-Sea Res., No. 4 (149-184)

Stommel, H. (1960) The Gulf Stream, University of California Press, Berkeley (202 pp.)

Sverdrup, H. U., M. W. Johnson and R. H. Fleming (1942) The Oceans, Their Physics, Chemistry and General Biology, Prentice-Hall, Inc.

Swallow, J. C. and B. V. Hamon (1960) Some measurements of deep currents in the Eastern N. A., Deep-Sea Res., Vol. 6 (155-160)

Tsuchiya, M. (1955) On a simple method of estimating the current velocity at the Equator, J. Oceanogr. Soc. Japan 11, (1-4)

Tsuchiya, M. (1955) On a simple method of estimating the current velocity at the Equator, Rec. of Oceangr. Works in Japan, 2, (1-6)

Veronis, G. (1956) Partition of energy between geostrophic and non-geostrophic motions, Deep-Sea Res. 3, (157-177)

Wüst, G. (1924) Florida-und Antillenstrom, eine hydrodynamische Untersuchung (Veröff Inst. f. Meereskunde, N.F. Reihe A, H. 12 Berlin (1-48)

Wüst, G. (1957) Stromgeschwindigkeiten und Strommengen in den Tiefen des Atlantischen Ozeans unter besonderer Berücksichtigung des Tiefen-und Bodenwassers. Wiss. Erg., Deutsch. Atl. Exped. "METEOR" 1925-27, Bd. VI, 2, Teil 6. Lfg.; (160) 6 abb., 36 Taf. u. 1 Beil, Berlin

Wyrtki, K., (1956) The computation of oceanic and meteorological fields of motion with friction proportional to the velocity, Penjelidkian Laut. di Indonesia, No. 2 (1-26)

Zubov, N.N. (1929) Computation of the elements of sea currents from hydrological profile data. Zap. po Gidrografii, V. 58

Zubov, N.N. and O.I. Mamayev (1956) Dynamic method of computing the elements of sea currents, GIMIZ

Zubov, N.N. (1957) Oceanological tables

Ocean Wave Measurements

Dr. Kenneth Hunkins
Wednesday, June 10, 1964

Introduction

Wave motions constantly disturb the surface and the density gradients of the oceans. These periodic motions are one of the important subjects of physical oceanography. Waves produce, primarily, a transfer of energy but not of matter. The ocean's wave motion is complex in space and time, involving wave lengths ranging from millimeters to thousands of kilometers and periods ranging from a fraction of a second to many days. Wave studies must begin with as accurate a description as possible of this motion within the limitations of instrumentation, their time and ingenuity.

I. Parameters which describe waves

A) These quantities may be determined by simple visual observation

1) C , Phase velocity

2) λ , Wave lengeh

3) T , Period

4) ψ , Direction of propagation

B) These parameters may be determined by visual observation and/or more exactly by recording instruments

1) $\zeta(x, y, t)$, Vertical displacement of the sea surface.

2) $\dot{\zeta}$, $\ddot{\zeta}$, Vertical velocity and vertical acceleration of the sea surface.

3) $\frac{\partial \zeta}{\partial x}$, $\frac{\partial \zeta}{\partial y}$, Slope of the sea surface

4) $P(z, t)$, Pressure at a given depth

5) (u, v, w), Oscillatory currents due to waves

C) These only apply to the case of waves from a limited source area; for example, from a seismic disturbance or from a distant storm

1) U , group velocity

2) $1/Q$, dissipation factor

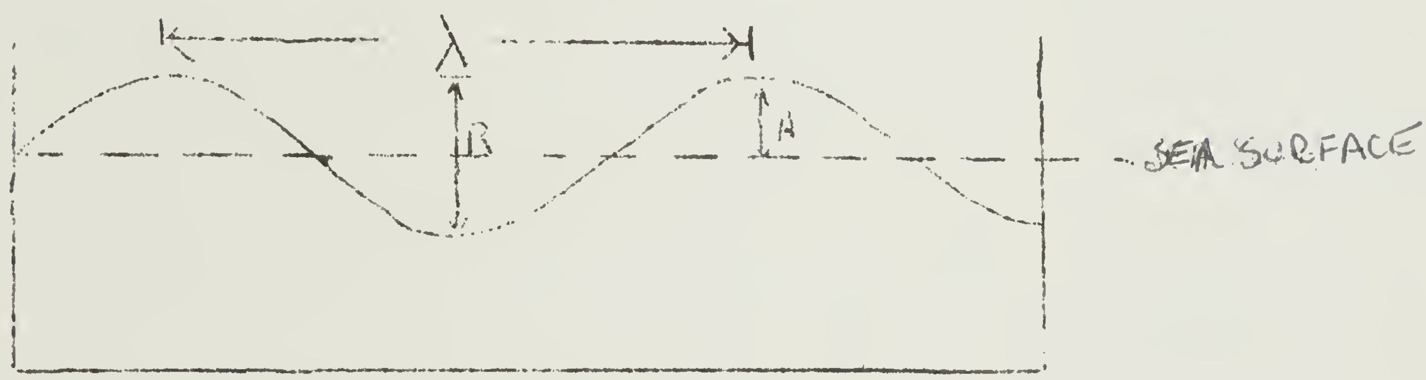
II. Ocean wave studies

A) To begin with, a simple harmonic motion is considered since the complex sea surface may be represented as a synthesis of sinusoidal wave trains with varying directions, amplitudes and periods. For a sine wave, the following definitions are customary:

1) Wave length - distance from crest to crest

2) Amplitude - displacement of disturbed sea surface at the crest from the undisturbed level.

3) Range - vertical distance from trough to crest; equal to twice the amplitude



B) For a freely progressing wave, the expression for the vertical displacement is

$$\xi = A \sin [ct - x] k$$

where ξ = vertical displacement

A = amplitude

k = wave number = $2\pi / \lambda$

x = horizontal distance

c = wave velocity

t = time

C) The hydrodynamic classification of waves is based on the physical parameters governing the motion, such as, restoring forces and boundary conditions.

1) For wave lengths greater than 1.7 cm., the restoring force is the earth's gravity. This wave length is the transition point between the short capillary waves and

longer gravity waves. The restoring force for the capillary waves is surface tension.

2) The gravity waves may be divided into two types, surface and long waves depending on the ratio of wave length to ocean depth.

3) Most of the energy is concentrated in the surface waves and they have received the most attention by investigators.

4) Information about the hydrodynamic properties of waves is summarized in the following table.

TABLE I

Wave Type	Capillary (Ripples)	Surface	Long
Restoring force	Surface tension	Gravity	Gravity
Particle orbit	Retrograde circles	Prograde circles	Linear
Phase velocity	$C^2 = kT$	$C^2 = g/k$	$C^2 = gh$
	where T = surface tension	where g = acceleration of gravity	where h = water depth
Dispersion	Inverse; group velocity greater than phase velocity	Normal; group velocity less than phase velocity.	No dispersion; velocity is independent of wave length

III. Wave recorders

A) Basic considerations in designing a wave recorder

1) Purpose of measurement and precision required

2) Frequency range to be covered and response curve of instrument

3) Reference system for measurements: ocean bottom, inertial mass, or for surface waves, the region of undisturbed water at depth.

B) Wave recorders used on board ship

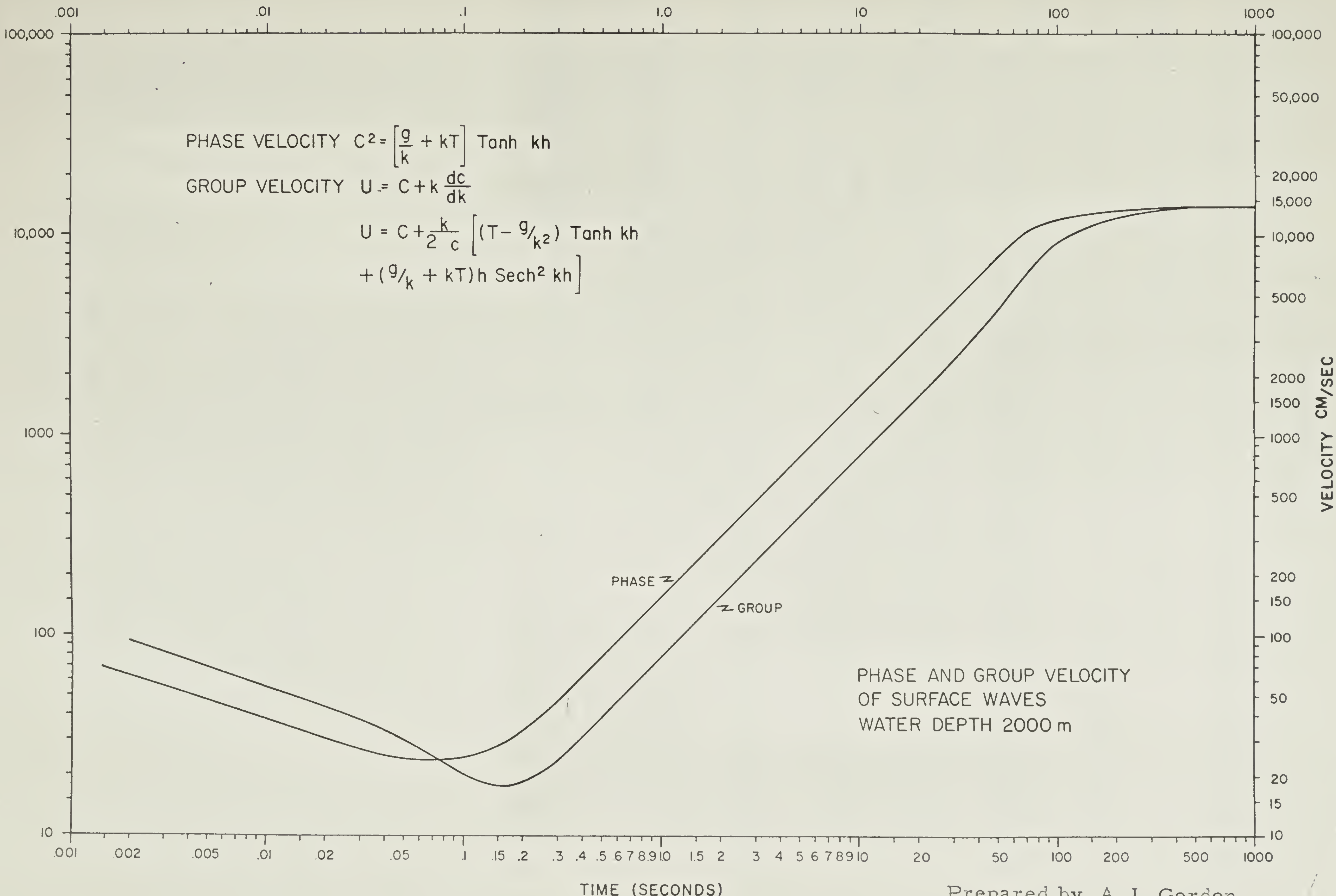
TABLE II

Type of recorder	Variable measured	Reference system	Filtering
Wave staff	$\zeta(t)$	Layer below wave action	Long wave filtering controlled by depth of suspension
Pressure recorder suspended below buoy	$\zeta(t)$	Layer below wave action	Long wave filtering controlled by depth of suspension
Surface buoy (N.I.O.)	$\zeta, \frac{\partial \zeta}{\partial x}, \frac{\partial \zeta}{\partial y}$	Inertial mass	Mechanical and electrical
Tucker ship recorder (N.I.O.)	$\zeta(t)$ through combination of vertical acceleration of ship with pressure on its hull	Ship's hull and inertial mass	Mechanical and electrical
Stereo cameras (Stereophotogrammetry)	Directional power spectrum	Ship's hull	

C) Wave recorder used at shore

TABLE III

Type of Instrument	Variable measured	Reference system	Filtering
Wave staff (mechanical, resistance or capacitance transducer)	$\zeta(t)$	Ocean bottom, staff is attached to dock or other structure	Hydrodynamic or electrical
Pressure recorder (Vibrotron transducer, Snodgrass; Capacitance transducer, Tucker)	$p(t)$	Ocean bottom	Hydrodynamic
Inverted echo sounder	$\zeta(t)$	Ocean bottom	Electrical
Tide gauge	$\zeta(t)$	Ocean bottom	Hydrodynamic
Pressure recorder (Munk)	$\zeta(t)$	Ocean bottom	Hydrodynamic and pneumatic



Prepared by A. L. Gordon

BIBLIOGRAPHY

Blackman, R. B. and J. W. Tukey (1958) The measurements of power spectra. Dover Publ., New York (190 pp.)

Coulson, C. A. (1952) Waves. Oliver and Boyd, Edinburgh

* Hill, M. N., ed., (1962) The Sea, Vol. I, Sect. V on Waves, Interscience Publishers

Munk, W. H., M.J. Tucker and F.E. Snodgrass (1957) Remarks on the ocean wave spectrum, Naval Hydrodynamics, Publ. 515, Natl. Acad. Sci.-Natl. Res. Council

Snodgrass, F. E., (1951) Wave recorders, Proc. 1st Conf. on Coastal Eng. Chap. 7, (69-81)

Tucker, M. J. (1955) A shipborne wave recorder, Proc. 1st Conf. Coastal Eng., Insts., Berkeley, Calif., Council on Wave Research, Univ. of Calif., Richmond, Calif.

Tucker, M.J. (1959) The accuracy of wave measurements made with vertical accelerometers, Deep-Sea Researcj. 5 (185-192)

"Ocean Wave Spectra" Proc. of Conf., Easton, Md., 1961. U.S. Natl. Acad. Sci., Prentice Hall, N. J. (1963)

* Particularly complete summaries of recent developments

Turbulence, Tides and Dynamical Oceanography

Dr. Takashi Ichiye
Thursday, June 11, 1964

Some topics in Dynamical Oceanography

Introduction

Although dynamic oceanography is a field of rather limited interest, understanding of its basic principles is essential for scientists in all branches of marine sciences. This is not only true for analyzing and interpreting the data obtained for different branches, but also for planning exploration, surveys or experiments in the ocean. Stommel (1963) illustrates this by considering two hypothetical survey plans: one for mapping deep water temperature to within ±0.05° C and the other for mapping month-by-month variation in sea level in the South Pacific within a significant figure of 2 cm. For the former plan, only 10^2 to 10^3 point observations taken over many years may be sufficient, while the latter plan requires at least 3×10^6 hourly observations at uniformly scattered stations for a period of several years when dynamical behaviors of the ocean as well as statistics of data are taken into account. Also, experiences in the recent international Indian Ocean Expedition indicate that a naive planning of hydrographic surveys without consideration of dynamics of ocean circulation may be almost useless in revealing a real situation.

This lecture will explain some basic principles and facts in dynamic oceanography which might be useful for practical oceanographers, in designing their own survey or experimental plan and in interpreting the data obtained.

I. Hydrodynamic Equations for the Ocean

A) Movement of the ocean water is governed by principles of hydrodynamics which are expressed with three sets of equations of motion and one equation of conservation of mass. Vectorial form of equations of motion is given by

$$\frac{D\vec{u}}{Dt} + 2\vec{\Omega} \times \vec{u} = -\frac{1}{\rho}\nabla p - \nabla G + \vec{F} \qquad (1)$$

in which $\vec{u}$, $\vec{\Omega}$ and $\vec{F}$ are vectors of velocity, rotation of the earth and frictional force, respectively; G is gravitational potential, ρ is density and $\frac{D(\ )}{Dt}$ is equal to $\frac{\partial(\ )}{\partial t} + \vec{u} \cdot (\ )$

B) The equation (1) represents Newton's law of dynamics applied to a fluid on the rotating earth. The first and second term represent an inertia and Coriolis' force, respectively and the third, fourth and fifth term represent the force due to pressure gradient, gravity force and frictional force, respectively.

C) Equations of conservation of mass and of any property of the water (temperature, salinity, dissolved oxygen, etc.) are expressed by

$$\frac{\partial \rho}{\partial t} + \nabla \cdot \rho \vec{u} = 0 \qquad (2)$$

$$\frac{\partial \rho q}{\partial t} + \nabla \cdot \rho q \vec{u} = - \nabla \cdot \vec{Q} \qquad (3)$$

respectively, where Q is the flux of q due to internal forces such as molecular heat conduction or molecular diffusion.

D) In order to see the importance of different terms in equation (1), velocity components, time and distances are expressed with non-dimensional quantities.

1) The x, y and z components are given by the following, where the positive directions of x, y and z are taken north, east and upward, respectively.

$$I \frac{\partial}{\partial t} u + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + \left(\frac{\sigma}{\delta}\right) w \frac{\partial u}{\partial z} - \frac{2}{R_o} \omega \sin\varphi\, v \qquad (4)$$
$$+ \frac{2\delta}{R_o} \omega \cos\varphi\, w = - \frac{1}{\rho} \frac{\partial p}{\partial x} + \frac{1}{R_e} \left(\nabla_h^2 u + \delta^{-2} \frac{\partial^2 u}{\partial z^2}\right)$$

$$I \frac{\partial}{\partial t} v + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + \left(\frac{\sigma}{\delta}\right) w \frac{\partial v}{\partial z} + \frac{2}{R_o} \omega \sin\varphi\, u \qquad (5)$$
$$= - \frac{1}{\rho} \frac{\partial p}{\partial y} + \frac{1}{R_e} \left(\nabla_h^2 v + \delta^{-2} \frac{\partial^2 v}{\partial z^2}\right)$$

$$I \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + \left(\frac{\sigma}{\delta}\right) w \frac{\partial w}{\partial z} + \frac{1}{R_o} 2\omega \cos\varphi\, u \qquad (6)$$
$$= - \frac{1}{\rho} \frac{\partial p}{\partial z} + \frac{1}{F_r} + \frac{1}{R_e} \left(\nabla^2 w + \delta^{-2} \frac{\partial^2 w}{\partial z^2}\right)$$

There are also six non-dimensional parameters which indicate relative importance of different terms:
$I = \frac{L}{U} T$, $R_o = \frac{U}{\Omega} L$ (Rossby number, $R_e = \frac{LU}{\nu}$ (Reynolds number; ν = molecular viscosity), $\delta = D/L$, $\sigma = W/U$, $F_r = WU/gL$ (Froude number).

The characteristic quantities are T for time; L for horizontal distance, D for vertical distance, U for horizontal velocity, W for vertical velocity, Ω for angular velocity of the earth and g for gravity constant.

2) For ordinary motion in the sea (excluding sound waves ripples, etc.), $F_r \ll 1$ and $R_e \gg 1$. The Rossby number R_o has wide variation for movement in the sea and ranges $R_o \gg 1$, $R_o \approx 1$ and $R_o \ll 1$ correspond to gravity waves, inertia-gravity motion and geostrophic motion, respectively. The motions corresponding roughly to the first and the third range was discussed by Dr. Hunkins and Mr. Gordon, respectively. Motions caused by tides and inertial oscillations belong to the second range.

II. Steady and Time-dependent Motion

A) The motion can be separated into steady and time-dependent motion for an adequate averaging time. When we separate the steady and time-dependent velocity as

$$\vec{u} = \vec{\bar{u}} + \vec{u'} \tag{7}$$

with the bar and dash indicating steady and time-dependent velocity, respectively, the equation of steady motion in x direction can be expressed in dimensional form by

$$\frac{\partial \bar{u}\bar{u}}{\partial x} + \frac{\partial \bar{u}\bar{v}}{\partial y} + \frac{\partial \bar{u}\bar{w}}{\partial z} - 2\Omega \sin\phi\, \bar{v} = -\frac{1}{\rho}\frac{\partial \bar{p}}{\partial x} - \left(\frac{\partial}{\partial x}\overline{u'u'} + \frac{\partial}{\partial y}\overline{u'v'} + \frac{\partial}{\partial z}\overline{u'w'} \right) \tag{8}$$

B) The Reynolds' stresses

1) In the above equation, the first, fifth and last terms of equation (4) are neglected. However, the last three terms of this equation are produced from the inertia terms and indicate the effect of time - dependent motion on the steady state motion. The tensorial form u'u' , u'v' is called Reynolds stresses. The main problems of hydrodynamics of turbulence is to determine relationships between Reynolds stresses and structure of main currents. In dynamic oceanography, too, these terms become important because they express energy exchange between the variable currents and the mean current. However, owing to lack of reliable instruments for measuring fluctuations of currents with sufficient accuracy and durability, our knowledge about oceanic turbulence is still insufficient.

2) In most papers dealing with the ocean currents, the Reynolds stresses have been expressed as similar to stresses due to molecular viscosity, using eddy viscosity instead of molecular viscosity as $u'w' = -A_z \frac{\partial \bar{u}}{\partial z}$, where A_z is vertical eddy viscosity. Recently such techniques as

self-recording current meters, neutral buoyant floats and dye diffusion have been utilized to determine dynamical structures of eddies which are predominant in the ocean.

BIBLIOGRAPHY

Books:

Defant, A. (1961) Physical Oceanography, Vol. 1, Pergamon Press, Chap. 9-13

Eckart, C. (1960) Hydrodynamics of Oceans and Atmosphere, Pergamon Press

Fofonoff, N. D. (1962) Dynamics of Ocean Currents, The Sea, Ed. M. N. Hill, Interscience Publishers, (323-395)

Townsend, A. A. (1956) The Structure of Turbulent Shear Flow, Cambridge University Press

Papers:

Phillips, N. A. (1963) Geostrophic Motion, Review of Geophysics 1, 2 (123-176)

Stommel, H. (1963) Varieties of Oceanographic Experience, Science 139 (572-576)

III. There are three approaches to study movement of water in the sea. One is the use of miniature models of oceans or limited bodies of water. The other is mathematical treatment of the equations of motion and continuity. The third is analysis of hydrographic data.

A) There are two kinds of models for simulating actual oceans or bays

1) Hydraulic models - Most of these models were constructed to study civil engineering problems in estuaries, bays or harbors, with emphasis on Froude number similarity. Rossby number of motion of the prototype is much less than unity in these models.

2) Rotating models- For the motion corresponding to the inertia-gravity and geostrophic ranges, the effect of Coriolis' terms cannot be neglected and a rotating model becomes necessary. Von Arx (1957) initiated experiments of the wind-driven circulation in the ocean with a rotating tank. Stommel, Arons and Faller (1958) utilized a rotating basin of a simple geometrical configuration such as a pie-shaped one to test the effect of northward increase of Coriolis' force on intensification of a western boundary current. Ichiye demonstrated that the rotation is essential in simulating the circulation in the medium-size sea, the Gulf of Mexico.

B) Mathematical models on oceanic circulation

1) Steady, wind-driven circulation

Stommel (1948) and Munk (1950) developed a theory on wind-driven ocean circulation, based on the vertically integrated equations of motion. The basic equations are:

(Vorticity Equation)

$$\beta V = \mathrm{Curl}\,(\vec{\tau} + \vec{F}) \tag{9}$$

(Continuity Equation)

$$\mathrm{div}\,\vec{U} = 0 \tag{10}$$

in which β is the latitudinal change of Coriolis' parameter, $\vec{U}$ is the vector of mass transport, τ is the wind stress and $\vec{F}$ is a frictional force. Assuming that the torque of frictional forces balances the absolute vorticity near the coast frictional boundary layer, they

could derive the most pronounced feature of the oceanic circulation; that is, existence of strong currents along the western coast of the ocean. Charney (1955) and Morgan (1958) discussed a non-frictional model and obtained the similar feature to the above model. They considered the torque of inertia forces balances the absolute vorticity (inertial boundary layer) near the coast. Greenspan (1962) derived the criterion of existence of the inertial boundary layer. Numerical integration by high-speed computers of non-linear equations of wind-driven circulation was recently started by several researchers (for example, Bryan, 1963) and indicates a great promise.

2) Steady state convective circulation

Fofonoff (1954) and Lineikin (1957) initiated a mathematical model study of ocean circulation due to convection with linearized equation of conservation of mass. Robinson and Stommel (1959), Welander (1959) and Robinson and Welander (1963) assumed the geostrophic flow with the non-linear equation of conservation of temperature. Their analysis is quite complicated even when the eddy diffusion is completely neglected or only vertical diffusion is retained. The main achievement of these studies is that they obtain the density stratification, particularly thermocline structure by a prescribed density distribution at the surface.

3) Time-dependent motion

Mathematical model of time-dependent oceanic circulation is less complete than the above two cases. The periodic (annual) change of wind-driven circulation of an enclosed ocean of uniform density was treated by Ichiye (1951) and Veronis and Morgan (1955). Veronis and Stommel (1956) treated the time-dependent motion in an unbounded ocean consisting of two layers. Free waves in such a system are classified into barotropic and baroclinic mode, according to whether both layers move in the similar way or only the upper layer moves. In each mode, there are gravity waves, inertia waves and Rossby waves, according to the relationships between the wave velocity and wave length (frequency equation) where gravity long waves are generally non-dispersive but the other two waves are dispersive. Ichiye (1958) developed a model of a two-layered system to the ocean bounded by east and west coasts.

C) Application of hydrodynamic principles to hydrographic data

1) It is very rare that hydrodynamic theories were applied to interpretation of hydrographic data except dynamic calculation of geostrophic flow. In particular, there are no data suitable to determine short-period (a few days to a few months) variation of ocean currents or energy interchange between the steady ocean currents and their perturbation, unlike the atmospheric circulation. This is mostly due to lack of frequencies in survey and poor design of survey plans, as Stommel (1963 loc. cit.) mentioned.

2) The principles which are most useful to application of hydrographic data are the conservation of potential vorticity

$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0 \tag{11}$$

and the conservation of mass

$$\operatorname{div}\left(H\vec{u}\right) = 0 \tag{12}$$

in which ζ and f are relative and planetary vorticity, respectively, H is the depth of the column of water. Rossby (1940) applied (11) to the meander of westerlies and to the forecasting of atmospheric jets (method of trajectories of constant vorticity). Later Ichiye (1955) and Saint-Guily (1957) treated the meander of the Kuroshio and the Gulf Stream, respectively, by applying Rossby's method to the ocean current. Warren (1962) and Greenspan (1963) incorporated the effects of bottom topography into (11) and (12) and indicated that the separation of the Gulf Stream and its subsequent meander pattern is due to the topography. However, recent survey of the Gulf Stream south of Grand Banks (Fuglister, 1963) seems to indicate that the absolute vorticity in the main part of the Gulf Stream might increase due to supply through Reynolds stresses.

3) It has been long suspected that eddies with scales of a few miles to 100 miles are a very common feature in the ocean. Reid (1963) for the first time determined velocity distributions in such an eddy (of about 30 miles diameter) off Southern California with multiple parachute drogues. It is speculated that this eddy was generated by a hydrodynamic instability in the shear zone between the northward coastal current and the southward offshore flow. (Pedlosky, 1963). Also, there is an implication that these intermediate-size eddies

may play an essential role in energy exchange between permanent durrents and the surrounding ocean water. It is hoped that more investigations on such eddies in other parts of oceans with drogues or anchored buoys will enhance our understanding on the maintenance and variation of oceanic general circulation

BIBLIOGRAPHY

Bryan, K. (1963) A numerical investigation of a non-linear model of a wind-driven ocean, J. of Atmosph. Sci. 20, 6, (594-606)

Charney, J. (1955) The Gulf Stream as an inertial boundary layer. Proc. Nat. Acad. Sci. 41, (731-740)

Fofonoff, N.P. (1954) A theoretical study of zonally uniform ocean flow. Tech. Rep. 4 for ONR of Brown University.

Fuglister, F.C. (1964) Gulf Stream'60 Progress in Oceanography, Vol. I, Editor - Mary Sears, (265-357)

Fultz, D. (1961) Development in Controlled Experiments on Larger Scale Geophysical Problems, Advances in Geophysics, Vol. 7, (1-104) Academic Press

Greenspan, H. P. (1962) A criterion for the existence of inertial boundary layers in oceanic circulation. Proc. Nat. Acad. Sci. 48 (2034-2039)

Greenspan, H. P. (1963) A note concerning topography and internal currents, J. Mar. Res. 21, 3 (147-154)

Ichiye, T.(1951) On the variation of oceanic circulation (1) Oceanog. Mag. 3, 2 (79-82)

Ichiye, T. (1955) On the variation of oceanic circulation in the adjacent seas of Japan. Proc. UNESCO Symp. on Oceanog. in 1955, Tokyo (116-129)

Ichiye, T. (1956) The response of a stratified bounded ocean to variable wind stresses, Oceanog. Mag. 10, 1 (19-63)

Lineikin, P.S. (1957) On the dynamics of the baroclinic layer in the ocean. Doklady Akad. Nauk. USSR 177, (971-974)

Morgan, G. W. (1956) On the wind-driven ocean circulation, Tellus 8, (301-332)

Pedlosky, J. (1963) Baroclinic instability in two-layer systems, Tellus, 15 (20-25)

Reid, J. L. (1963) Direct measurements of a small surface eddy off Northern Baja California, J. Mar. Res. 21, 3 (205-218)

Robinson, A. and H. Stommel, (1959) The oceanic thermocline and the associated thermo-haline circulation Tellus 11, (295-308)

Robinson, A. and P. Welander (1963) Thermal circulation on a rotating sphere with application to the oceanic thermocline, J. Mar. Res. 21, 1 (25-38)

Rossby, C. G. (1940) Planetary flow patterns in the atmosphere. Suppl. Quart. J. Roy. Meteorol. Soc., (68-87)

Saint-Guily (1957) Les méandres des veines de courant dans les océans. Bull. l'Inst. Oceanog. Monaco 1108 (1-11)

Stommel, H., A. B. Arons and A. J. Faller (1958) Some examples of stationary planetary flow patterns in bounded basins, Tellus 10, (179-187)

Veronis, G. and G. W. Morgan (1955) A study of the time-dependent wind-driven ocean circulation. Tellus 7, 2 (232-242)

Veronis G. and H. Stommel (1956) The action of variable wind stresses on a stratified ocean, J. Mar. Res. 15, 1 (43-75)

Von Arx, W. S. (1957) An experimental approach to problems in physical oceanography. Progress in Physics and Chemistry of the Earth, 2, (1-29) Pergamon Press

Warren, B. A. (1963) Topographic influences on the path of the Gulf Stream, Tellus 15, 2 (167-183)

Welander, P. (1959) An advective model of the ocean thermohaline, Tellus 11 (309-318)

Topics in Chemical Oceanography

Friday, June 12, 1964

Salinometry and Saturometry

Willard Moore

I. "A salinometer may be defined rather freely as an instrument which determines the salinity of sea water by measuring some property of the water which has a constant relation to the salinity." (Paquette, 1959).

A) Salinity is defined as the total amount of solid material in grams in one kilogram of sea water after all carbonate has been converted to oxides, all bromine and iodine replaced by chlorine, and all organic material oxidized. (Forch, Sorensen and Knudsen, 1902)

B) The classical method of measurement is by a chemical titration with $AgNO_3$ (See R. Sexton's section) to find the amount of chlorine present. The chlorinity is converted to salinity by the imperical formula

Salinity = 0.03 + 1.805 (chlorinity)

The accuracy of this method is no better than ± 0.01 o/oo

C) Use of conductivity

1) The electrical conductivity can be measured and this converted to chlorinity. This method measures the ionic strength of the solution. It is extremely sensitive to temperature, and a salinometer must be equipped with a temperature compensation device. The relation of conductivity to chlorinity is not known exactly. The predicted relationship is linear (i.e. conductivity raising linearly with chlorinity at any one temperature); however, surface samples fall below the line and deep water samples are above the predicted line. This method is, on the whole, more accurate than the chemical method; most oceanographers give its accuracy to ± 0.003 o/oo in salinity, but there is doubt to this figure.

2) The relationship of conductivity to density is said to be more accurately known than conductivity to chlorinity and if it is density which is needed, it may be best to go directly to this relation. This would not be useful until in situ measurement of conductivity is perfected. Since density is not

always the prime reason for measuring salinity (core method), the true relation of conductivity to chlorinity must be known. This problem is being investigated by various oceanographers.

D) All salinometers must be calibrated with standard Copenhagen sea water; this is water of known chlorinity. A study of the reliability of standard water indicates that a standard deviation of .001% may be expected. The recommendation has been made that standard water be certified for its conductivity rather than for chlorinity

II. Saturometry is the measurement of the degree of saturation of $Ca\,CO_3$ in the sea water.

A) Method of measurement depends on the pH change as $CaCO_3$ is dissolved or precipitated. If there is no change when one of the pH determining electrodes is brought into contact with solid $CaCO_3$, the solution is saturated. If the pH goes down this indicates that $CaCO_3$ is being precipitated and releasing hydrogen ions; therefore, the solution was supersaturated. To get quantitative results, the saturometer must be calibrated with standard sea water or titrated with $Na_2\,CO_3$ solution to same pH change.

B) Importance - There is disagreement as to the relative influence that temperature and pressure have on the resolution of $CaCO_3$ from sediments. A temperature effect has been observed with sediments (i.e. below cold water masses, the $CaCO_3$ in the salinometer is generally depleated); however, calculations would indicate that pressure was the more important. By controlled experiments, the saturometer may resolve this discussion.

Special Topics in Marine Chemistry	Given by
Water Sample Collection and Marine Preservation	Dr. W. Sackett
Use of Oxygen Isotopic Ratio in Oceanography	Dr. W. Sackett
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Bibliography

Broecker, W., (1963) Radioisotopes and large-scale oceanic mixing, The Sea, Vol. 2 (88-108)

Craig, H., Gordon, L.I. and Horibe, Y., Isotopic Oceanography, Scripps Inst. Ocean. ONR Report 1962-63

Paquette, R. G., (1959) Salinometers, Physical and Chemical Properties of Sea Water, Nat. Acad. Sci. Pub. 600 (128-145

Weyl, R. G., (1961) The carbonate saturometer, Journ. Geology 69, (32-44